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WEIGHTING FUNCTION ANALYSIS FOR THE PENN STATE UPPER ATMOSPHERE--ETC(U)
JAN 81 R M BEVILACQUA N00014-79-C-0610
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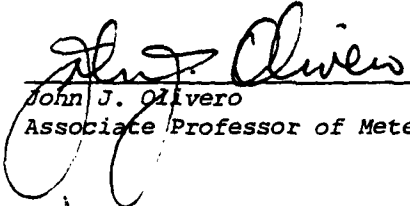
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Abstract

The measurement of water vapor in the middle atmosphere by microwave radiometry, using the 22.235 GHz water vapor absorption line, is being pursued at the Ionosphere Research Laboratory and the Department of Meteorology here at Penn State. In this report the weighting functions which are required in order to invert the radiance data to obtain the corresponding water vapor profile are derived and calculated. An examination of these weighting functions indicates that, with our instrument, water vapor measurements can successfully be made in the 50 to 85 km altitude range.

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1. Introduction

The measurement of water vapor in the middle atmosphere by microwave radiometry is being pursued at the Ionosphere Research Laboratory and in the Department of Meteorology. In brief the goal of the project is to use our ground-based radiometer, which is presently nearly completed, to produce a spectrum of radiance measurements across the 22.235 GHz water vapor absorption line, and to use this data to infer the mesospheric water vapor profile. It is hoped that once the experimental program is established, mesospheric water vapor will be monitored on a continuous basis. The details of the instrument are discussed elsewhere. Here it need only be mentioned that spectral analysis of the radiometer data will be obtained with the use of a filter-bank spectrometer which comprises 50 contiguous channels each of 50 KHz bandwidth for a total bandwidth of 2.5 MHz, centered at line center. In this report the weighting functions which are required in order to retrieve a water vapor profile from the radiance measurements are derived. An examination of these weighting functions demonstrates the feasibility of measuring mesospheric water vapor with our instrument.

2. Weighting Functions

In general the weighting function for any type of remote sensing experiment can be defined such that

$$g(y) = \int WF(x,y) f(x) dx \quad (1)$$

where $g(y)$ is the measured quantity,

$f(x)$ is the quantity to be inferred from the measurement

$WF(x,y)$ is the weighting function for the particular experiment.

In our experiment the vector $g(y)$ is the radiance measurement, with y being the frequency at which an individual measurement is made. The vector $f(x)$ becomes $n_{H_2O}(s)$, the water vapor profile as a function of height. In order to determine the manner in which radiance measurements in the vicinity of the 22.235 GHz water vapor absorption line are related to the water vapor profile, and thus determine the weighting function, an equation of radiative transfer must be used. The radiative transfer equation, for an upward viewing path through the atmosphere, applied at microwave frequencies, where atmospheric scattering can be neglected and $h\nu \ll kT$ (long wavelength limit of the Planck function), takes the simple form (Waters, 1976).

$$T_g(\nu) = T_\infty(\nu)e^{-\tau(\nu, \infty, 0)} + \int_0^\infty T(s)e^{-\tau(\nu, s, 0)} K_a(\nu, s) ds \quad (2)$$

where $T_g(\nu)$ is the brightness temperature, at frequency ν , at the ground,
 $T_\infty(\nu)$ is the brightness temperature of any microwave source external to the atmosphere,

$T(s)$ is the atmospheric temperature profile,

$K_a(\nu, s)$ is the volumetric absorption coefficient and

$\tau(\nu, s, 0)$ is the atmospheric optical depth or opacity,

$$\tau(\nu, s, 0) = \int_0^s K_a(\nu, s') ds'.$$

Two types of experiments can be performed and simulated using Eq. (2): an absorption experiment in which the radiometer's antenna is mounted on a solar tracker, and the attenuation of solar radiation is measured; and an emission experiment in which the antenna is fixed and the thermal radiation emitted by the atmosphere is measured. The relative merits and disadvantages of each type of experiment are discussed in detail in Longbothum (1976), and the interested reader is

referred there. In the following subsections the appropriate weighting function for each experiment will be derived.

a) Absorption Experiment

When looking at the sun the radiation emitted by the atmosphere can be neglected because it is extremely small by comparison, therefore the radiative transfer equation reduces to:

$$T_g(\nu) = T_{\text{sun}}(\nu) e^{-\tau(\nu, \infty, 0)} \quad (3)$$

Equation (3) can be rewritten as

$$\ln \left(\frac{T_{\text{sun}}(\nu)}{T_g(\nu)} \right) = \tau(\nu, \infty, 0) = \int_0^{\infty} K_a(\nu, s) ds.$$

In the stratosphere and mesosphere the absorption coefficient for frequencies within a few MHz of the 22.235 GHz water vapor absorption line is, to a good approximation, linear in the water vapor concentration and can be written

$$K_a(\nu, s) = \sigma(\nu, s) n_{\text{H}_2\text{O}}(s) \quad (4)$$

where $n_{\text{H}_2\text{O}}$ is the water vapor concentration,

$\sigma(\nu, s)$ is the water vapor absorption cross section for the 22.235 GHz line.

With this approximation Eq. (3) can be written in its final form

$$\ln \left(\frac{T_{\text{sun}}(\nu)}{T_g(\nu)} \right) = \int_0^{\infty} WF(\nu, s) n_{\text{H}_2\text{O}}(s) ds \quad (5)$$

where $WF(\nu, s)$ is simply equal to $\sigma(\nu, s)$.

The left side of Eq. (5) can be considered known from spectral analysis of the radiometer data. A potential problem with this is that the solar brightness temperature can be quite variable at

microwave frequencies, hence T_{sun} is not well known. This problem can be circumvented by performing the measurement at more than one zenith angle, and thus eliminating T_{sun} from Eq. (5).

In order to calculate the weighting functions, for the frequencies which correspond to the filter bank used in our measurement program, the microwave radiative transfer model developed by Longbothum (1976) has been used. This model was developed in conjunction with a feasibility study, carried out at the Ionosphere Research Laboratory, of using microwave radiometry to measure water vapor in the stratosphere and mesosphere. In Figure (1) we present a plot of the absorption weighting functions for the first 25 channels of the filter bank, each normalized by its maximum value. Thus the curves are separated by 50 KHz with the offset from line center ranging from 1.25 MHz to zero. The line is symmetrical about line center so that the 25 channels on the opposite side of line center give essentially redundant information. The line is centered on the center channel of the filter bank to produce a symmetric spectrum, in order to avoid any problems with baseline instabilities. Figure (1) illustrates that the weighting functions peak at successively higher altitudes with decreasing frequency offset in an apparently exponential fashion. The figure also shows that these weighting functions cover the 50 to 85 km altitude range quite well. Therefore water vapor profile measurements in this altitude range are feasible with our instrumentation. One other point bears mentioning, with the 50 KHz resolution there is a considerable amount of overlap,

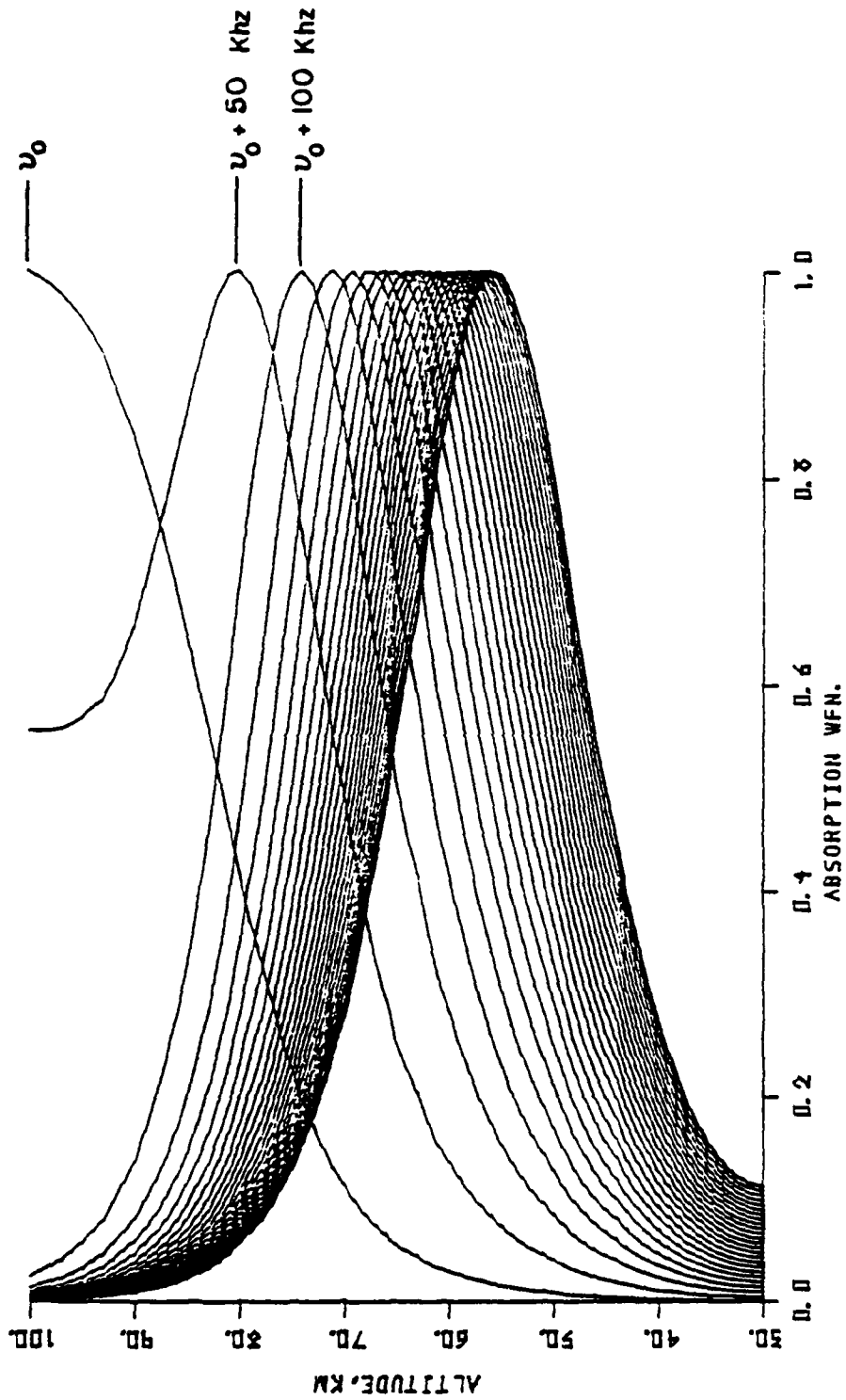


Fig. 1: Normalized absorption experiment weighting functions for the first 25 contiguous filter bank channels, with $\nu_0 = 22.235 \text{ GHz}$.

and therefore redundancy of information, in the weighting functions. In an actual data inversion procedure not all these frequencies would be included. Another scientific report objectively determines the number of independent pieces of information, and thus the frequencies which should be used in the data inversion, based on the experimental error in the measurements.

b) Emission Experiment

In an emission experiment the antenna is pointed in a direction away from the sun in order to measure the thermal radiation emitted by the atmosphere. In this case the radiative transfer equation becomes

$$T_g(\nu) = \int_0^{\infty} T(s) e^{-\tau(\nu, s, 0)} K_a(\nu, s) ds. \quad (6)$$

By definition, then, the weighting function for this particular experiment is given by

$$WF(\nu, s) = T(s) e^{-\tau(\nu, s, 0)} K_a(\nu, s) / n_{H_2O}(s) = T(s) e^{-\tau(\nu, s, 0)} g(\nu, s)$$

so that Eq. (6) can be written

$$T_g(\nu) = \int_0^{\infty} WF(\nu, s) n_{H_2O}(s) ds.$$

At first glance this expression for the weighting function appears to be of little use because it contains the opacity for the underlying atmosphere, $\tau(\nu, s, 0) = \int_0^s K_a(\nu, s') ds'$, which depends on the water vapor profile from the given level, s , down to the surface. However, the atmospheric opacity is dominated by tropospheric attenuation. In fact the Longbothum model, with a reasonable water vapor profile, shows that the opacity from the tropopause

to the surface is more than 98% of the total opacity from the surface to 100 km. Thus in the stratosphere and mesosphere the atmospheric opacity can be considered essentially independent of height, and the emission weighting function can be written in the final form

$$WF(\nu, s) = T(s)\sigma(\nu, s)e^{-\tau(\text{trop})} \quad (7)$$

It can be seen from Eq. (7) that the emission weighting function differs from its absorption counterpart only in that it has an extra linear temperature dependence. However, the absorption cross section varies over orders of magnitude in the mesosphere, while the temperature varies over tens of percent. Thus, variations in the emission weighting function will be controlled mainly by variations in $\sigma(\nu, s)$. In Figure (2) the emission weighting functions are plotted in the same format as were the absorption weighting functions. As expected no significant differences between the two types of weighting functions are observed. Hence, the same altitude range can be covered with the absorption and the emission experiment.

The weighting functions can be carried one step further to determine, in general, the vertical range of water vapor profile measurements, using ground-based microwave techniques, as a function of the instrument bandwidth. The highest altitude at which water vapor measurements can be made using the 22.235 GHz absorption line, occurs at approximately 85 km. This is a consequence of the fact that above that altitude Doppler broadening begins to surpass pressure broadening as the dominant line broadening mechanism, and the line width is no longer a function of pressure, or altitude.

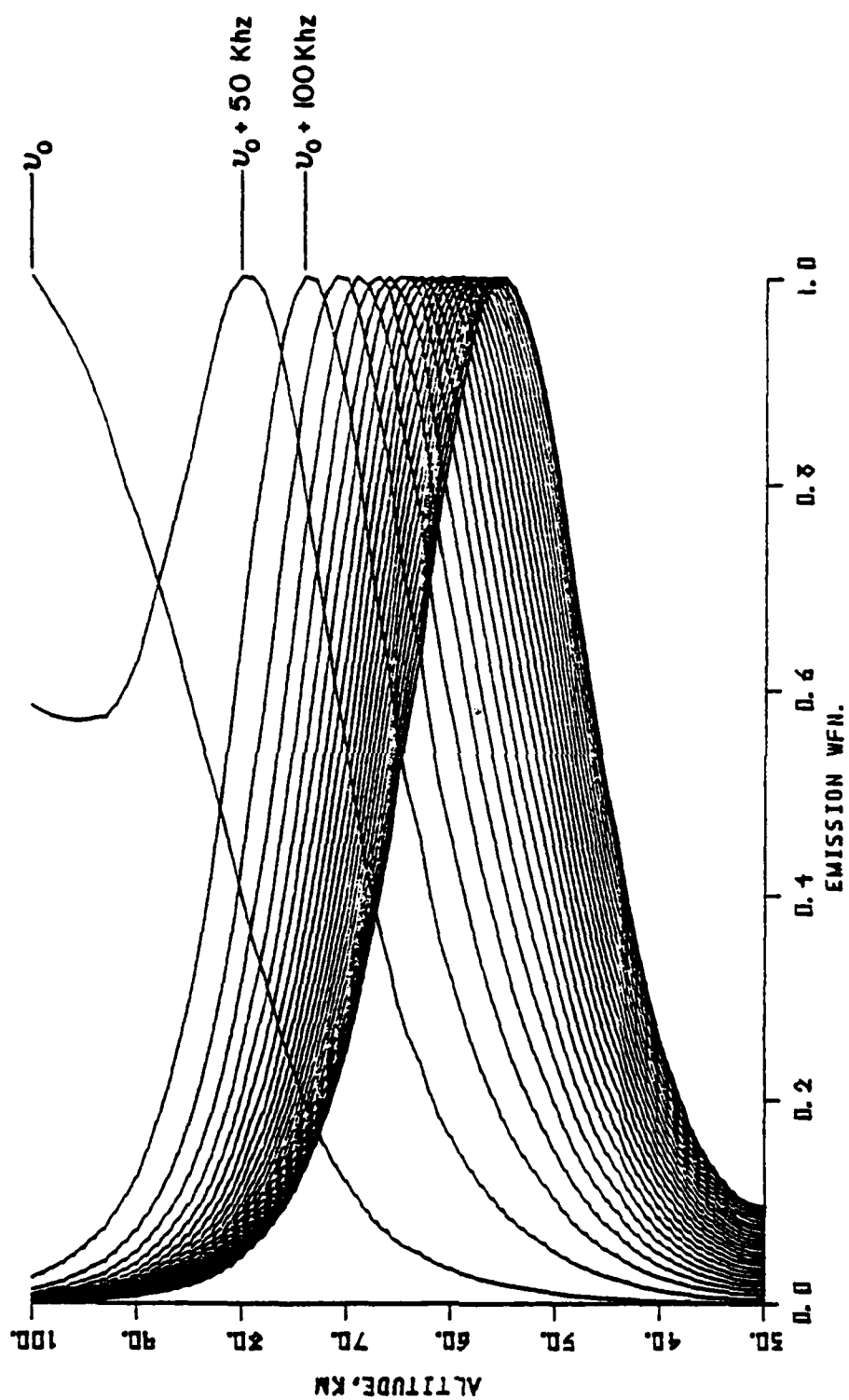


Fig. 2: Normalized emission experiment weighting functions for the first 25 contiguous filter bank channels, with $\nu_0 = 22.235 \text{ GHz}$.

The lowest measurement altitude for a given instrument can be estimated as the altitude corresponding to the weighting function peak, at maximum frequency offset. Figure (3) is a plot of the altitude of the peak of the weighting function as a function of frequency offset. The exponential nature of the weighting functions, which was mentioned previously, is evident in Figure (3). Because of this, it is extremely difficult to employ ground-based microwave techniques to make mid and lower stratospheric water vapor measurements. For example, to obtain water vapor profiles successfully down to 20 km would require an instrument with a bandwidth of greater than 100 MHz. The large bandwidth involved, then, tends to limit the vertical range of measurements using this technique, because wide bandwidth spectrometers suffer from baseline instability problems. The weighting functions, however, remain well behaved into the troposphere; therefore, in principle measurements can be made down into the troposphere, given an instrument of sufficiently large bandwidth, and good baseline stability.

3. Conclusion

Emission and absorption weighting functions have been calculated for the frequencies which correspond to those obtained from spectral analysis of the radiometer data, with the filter bank spectrometer. An examination of these weighting functions reveals that it is quite feasible to make mesospheric water vapor measurements with our present instrumentation in both the absorption and emission modes.

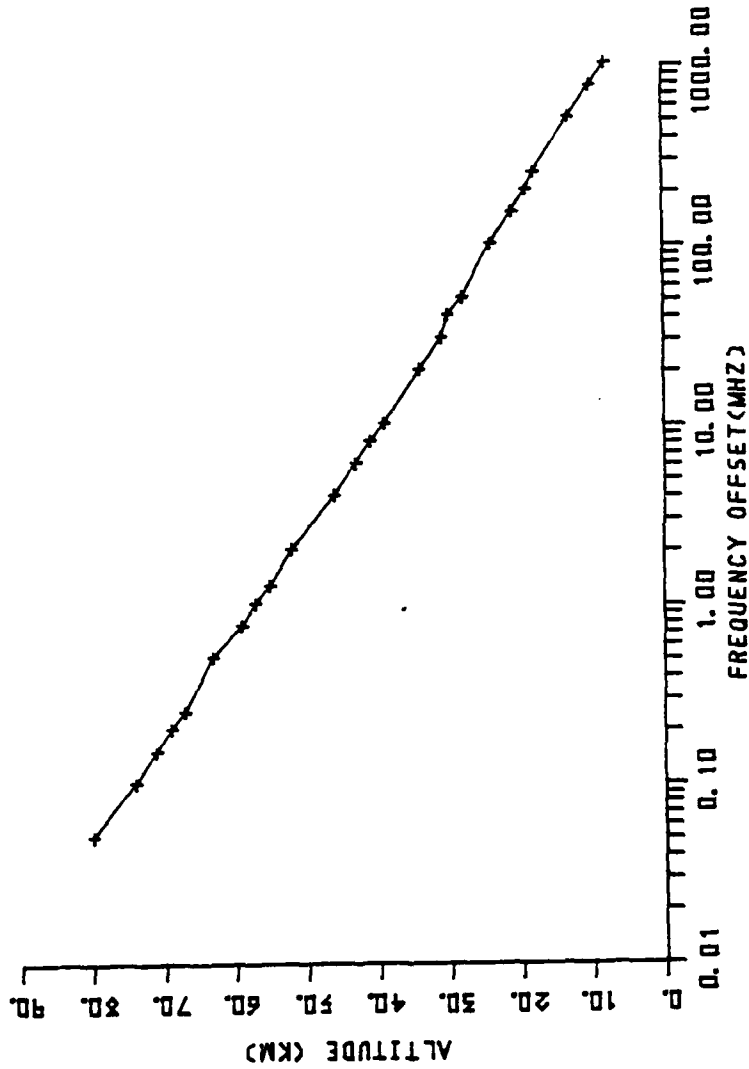


Fig. 3: Altitude of weighting function peak, as a function of frequency offset from line center for the 22.235 GHz water vapor line.

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The measurement of water vapor in the middle atmosphere by microwave radiometry, using the 22.235 GHz water vapor absorption line, is being pursued at the Ionosphere Research Laboratory and the Department of Meteorology here at Penn State. In this report the weighting functions which are required in order to invert the radiance data to obtain the corresponding water vapor profile are derived and calculated. An examination of these weighting functions indicates that, with our instrument, water vapor measurements can successfully be made in the 50 to 85 km altitude range.

PSI-IRL-SRI-489
Classification Numbers:

1.9 Structure of the Upper Atmosphere

1.2.1 Ground-Based Techniques and Measurements

1.2.3 Data Analysis Techniques

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